THE CRUST OF VENUS

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Surface geochemical measurements and the morphology of volcanic structures imaged from orbit indicate that the surface of Venus is dominated by basalt. Even unusual geochemical signatures and volcanic landforms suggestive of silicic lavas are consistent with mafic to intermediate bulk compositions. Gravity data indicate that the mean thickness of the crust is 20 to 50 km; 30 km is the value preferred here. The planet's smooth hypsometry and the dominance of deep isostatic compensation argue that the crust has a comparatively uniform thickness, departing significantly only in the tessera highlands. The crust at present occupies 1 to 2% of the total volume of Venus. By contrast, Earth's present fractional volume of basaltic crust is ~0.5%, but may integrate to ~10% over the last 4 Gyr due to crustal recycling. In spite of the recent evidence from the Magellan mission that Venus underwent strong volcanotectonic resurfacing at or before several hundred million years ago, it is still unknown whether such resurfacing included significant crustal recycling. Recycling is required if the bulk of the present crust was generated by horizontal accretion analogous to seafloor spreading. However, full development of horizontal crustal accretion is difficult to envision because of the depth and temperature of melting necessary to generate basaltic crust so much thicker than terrestrial oceanic crust. In addition, the bulk composition of the crust must be picritic to komatiitic and a late global flooding event must conceal features diagnostic of earlier lithospheric recycling. Alternatively, a dry, stiff upper mantle may have allowed only limited horizontal movements, and vertical accretion dominated crustal generation. A tholeiitic to picritic crust could be formed at numerous regional centers of lithospheric extension during episodes of global activity. During periods of global quiescence, alkali basalts would be a significant melt component produced by minor stretching of a thick lithosphere or plume melting beneath a rigid lid. Recycling of crust is possible but not necessary in this model. The present crust is probably too thin to be globally limited by densification and detachment due to the granulite/eclogite phase transitions, but such controls, especially locally, cannot be ruled out. Efficient distribution of magma or subsequent solid-state creep is necessary to inhibit large permanent lateral variations in crustal thickness under vertical accretion. Contemporary crustal production is 0.01 to 0.4 $km^3 yr^{-1}$, less than or equal to the intraplate crustal formation rate of Earth.

I. INTRODUCTION

The crust of a moon or planet is a thin outer layer that differs markedly in composition from the interior and from the composition of the primordial solar nebula (Taylor 1989). It further concentrates sizeable fractions of the planetary budget of incompatible elements. The crust is a chemical boundary layer formed by internal differentiation under a variety of melting processes and is distinct from the mechanical boundary layer or lithosphere. Because of the complexity of silicate phase relations, a range of crustal mineralogies is produced on rocky worlds. No such distinctions have yet been demonstrated for the icy moons.

Due to its similar bulk properties with Earth, Venus has been expected to show evidence for a similar thermal history and perhaps a comparable record of crustal differentiation. A variety of experimental and theoretical efforts based on Pioneer Venus (PV), Venera, and Vega orbital and surface data attempted to constrain the composition, volume, and differentiation history of the crust. Geochemical measurements made by a series of landers between 1973 and 1986 were largely consistent with a basaltic surface (Barsukov 1992). In the first Venus volume, McGill et al. (1982) explored differences in the rock cycle and hypsograms between Earth and Venus. They further determined that plate boundaries should be visible on Venus if present. Grimm and Solomon (1988) used geodynamic methods to determine the crustal thickness and inferred that the total crustal volumes of Earth and Venus were similar. They concluded that either Venus generated crust slowly or that some form of crustal recycling had occurred. Hess and Head (1990) discussed the compositions expected from a variety of crustal differentiation mechanisms. They found that, in spite of the hypothesized lack of water in the crust and mantle, no major Earth-like igneous suites could be excluded, although some systematic differences should be present. Head (1990) summarized many of these pre-Magellan ideas.

Perhaps the most startling result from Magellan is the implication from the cratering record that Venus underwent rapid resurfacing at or before several hundred million years ago, which then sharply declined (Schaber et al. 1992). Geodynamic models seek to understand the internal processes responsible for this abrupt transition, and to determine whether such changes are monotonic or episodic (Arkani-Hamed et al. 1993; Turcotte 1993; Parmentier and Hess 1992; chapter by Schubert et al.). Although Magellan gathered no primary geochemical data, these results have profoundly influenced geophysical concepts on the mechanisms, timing, and rate of crustal formation. In particular, the abandonment of a strictly uniformitarian perspective means that contemporary crustal production and recycling processes may be different, or least operate at a much reduced rate, than in the past.

In this chapter, we examine the crust of Venus in light of the Magellan results. As just mentioned, these data have a greater geophysical than geochemical impact, so our emphasis follows. We first review the mechanisms of crustal formation, both for terrestrial planets in general and Venus in particular. We re-assess the different kinds of evidence for crustal compositions on Venus. We analyze several methods for determining the crustal thickness (and volume) of the planet, and compare to the Earth. Lastly, we present three alternative hypotheses for crustal production and recycling on Venus in the past and estimate the recent rate of crustal formation.

II. CRUSTAL FORMATION ON THE TERRESTRIAL PLANETS

Taylor (1989) classified planetary crusts into three broad categories. A primary crust is formed by nearly complete melting and then subsequent crystal fractionation of the upper mantle during the late stages of accretion. For the Moon, the primary crust consists mostly of anorthosite which, according to the magma ocean hypothesis, formed by the flotation of calcic plagioclase over a global magma system (see Warren and Kallemeyn 1993, for an update). It is generally agreed that the magma system must have been large, as the Al_2O_3 content residing in the primary crust probably represents >70% of the global inventory (Wood 1986; O'Neill 1991). But how likely is it that the Earth or Venus once developed a stable primary crust? Taylor (1989) pointed out several reasons why Earth probably lacked an early anorthositic crust. The most relevant of these to the ensuing discussion of Venus is that plagioclase is not a liquidus phase in basalt melts above $\sim$1 GPa. These pressures correspond to depths of <40 km on Venus and Earth but up to 200 km on the Moon. Thus, the depth range on Venus and Earth over which to form stable anorthosite crust by crystal flotation is very limited. Primary anorthositic crust has been confirmed only for the Moon. Such a crust is likely to have formed on Mercury and has been inferred from spectroscopic measurements (Jeanloz et al. 1995). At present there is no evidence for primary crust on Mars; spectroscopic observations suggest Mars is dominated by oxidized basalts and other alteration products (Soderblom 1992).

Secondary crusts are the result of regional, rather than global, melting and differentiation of the upper mantle. For silicate planets, this crust is overwhelmingly basaltic for the simple reason that basalt is the product of pressure-release partial melting of peridotite, which is expected to dominate the bulk composition of the mantles of the terrestrial planets. The exact composition is a function of the temperature at which melting is initiated and the degree of partial melting. Pressure-release melting is a consequence of mantle upwelling and/or lithospheric stretching. Terrestrial seafloor spreading forms prodigious quantities of secondary crust. Furthermore, seismic constraints on the continental crust (Christensen and Mooney 1995) and mass-balance calculations for island arcs (DeBari and Sleep 1991) indicate that upper intermediate-to-silicic rocks are in both cases underlain by mafic rocks, which could represent additional formation of secondary crust. Sample analyses have confirmed the presence of basaltic secondary crusts on the Moon, Venus, and Mars; the morphology of plains regions on Mercury suggests that low-viscosity basalts may be present there also (Spudis and Guest 1988).

Tertiary crusts are formed by remelting pre-existing crust. Most terrestrial tertiary crust is formed at convergent margins where melting of the peridotite mantle wedge is initiated by the release of H_2O by dehydration of metamorphosed oceanic basaltic crust (Hess 1989). Underplating and intrusion of continental crust on Earth by basaltic magma has also been invoked as a mechanism to initiate magmatism in a variety of tectonic settings (Hildreth and Moorbath 1988). The remelting of subducted basaltic crust itself is a more contentious issue, although generation of trondhjemite-tonalite-dacite melts from relatively young subducted plates seems to be a reasonable hypothesis (Drummond and Defant 1990). Nonetheless, the key to the formation of andesitic tertiary crust on Earth is the ability to recycle H_2O-rich volatiles into the mantle. The existence of large volumes of tertiary crust on Venus and, indeed on other planets, will most likely depend on the same variables.

III. EXPECTED CRUSTAL COMPOSITIONS ON VENUS

Hess and Head (1990) gave a comprehensive summary of the petrogenetic constraints on crustal formation on Venus. Here we briefly summarize and expand upon this work, reorganizing the discussion to follow more closely Taylor's framework outlined above.

Venus does not display any elevated regions of distinctly higher crater density analogous to the lunar highlands that may indicate primary crust preserved since just after planetary formation. Furthermore, the young average impact crater retention age—comparable to the average age of the surface of the Earth—indicates a vigorous geological history. Therefore it is probable that primary crust has been completely recycled sometime over the last 4 Gyr. Alternatively, stable primary crust may never have formed on Venus, as Taylor argues for the Earth. However, we note that the present evidence cannot completely exclude pockets of primary crust; direct geochemical measurements are sparse, and crater retention ages do not correspond to crystallization ages where the surface has been strongly reworked, as in the highly deformed "complex ridged terrain" or tessera. We are not suggesting that tessera actually are primary crust, merely that crater retention ages are not indicators of crustal composition.

Bulk density and cosmogenic arguments predict that the mantle composition of Venus should be broadly similar to the Earth's (BVSP 1981) and hence the mantle should be peridotitic. Basalts are therefore expected as primary regional mantle melts forming secondary crust. A key parameter controlling the composition and amount of melt is the mantle potential temperature T_p, defined as the temperature that adiabatically upwelling material would reach at the planetary surface if it did not melt (see, e.g., McKenzie and Bickle 1988). The potential temperature is constant along an adiabat, whereas the actual temperature varies. Depending on T_p and the amount of upwelling, the full range of composition from quartz tholeiites ($T_p \sim 1400$ K) to komatiites ($T_p \sim 1800$ K) can be produced. The presence of volatiles, however, has significant effects on the melting regime. The role of H_2O is not only to lower the solidus temperature, but also to make near-solidus melts less olivine-normative compared to the dry system. The presence of CO_2 has the most dramatic effect at pressures greater than about 2 GPa. Near-solidus melts are more olivine-normative in a CO_2 atmosphere compared to the dry system. At very small degrees of melting, strongly SiO_2-undersaturated melts ranging in composition from alkali basalts to kimberlites and even carbonatites can be produced. Kaula (1990,1993) has argued that the lack of a moon-forming impact on Venus may have left it with an appreciable volatile content in the lower mantle.

Recycling of secondary basaltic crust can produce a range of tertiary crusts depending on the depth of melting and the volatile content. For shallow underthrusting and remelting of basalt under dry reducing conditions, ferrobasalt melts are expected for modest degrees of melting (pressures less than about 1.0 to 1.5 GPa and melt fractions greater than about 0.1). Silica-rich melts could result at smaller degrees of melting but these would be difficult to erupt. Such melts can be obtained at larger degrees of melting of basalt under more oxidizing conditions, however. At greater depths where basalt converts to eclogite ($P > 1.5$ GPa for most basalt compositions), more silica-rich melts such as trondjhemites, andesites, or even dacites are formed (Johnston 1986). Melting tholeiite basalt with H_2O or even a CO_2-H_2O fluid causes the solidus to be depressed by several hundreds of degrees and generates near-solidus SiO_2-enriched melts. Baker and Eggler (1987), for example, produced dacite melt with CO_2-H_2O fluids containing less than 25% H_2O at pressures less than 0.8 GPa. The effects of melting basalt in a pure CO_2 atmosphere are not well constrained, however. See Hess and Head (1990) for a more extensive discussion on the origin of silicic magmas.

IV. INFERENCES OF CRUSTAL COMPOSITION

There are two broad classes of evidence that have been used as indicators of crustal compositions on Venus. The first consists of direct measurements of elemental or oxide abundances from the Venera landers. The second class consists of indirect inferences from geomorphology, particularly the characteristics of lava flows. We will review these approaches in detail, as well as other suggestions derived from gravity and topography.

A. Lander Measurements

The most direct evidence for the compositions of crustal rocks on Venus has been provided by seven spacecraft with geochemical instruments that were successfully landed on Venus by the former Soviet Union between 1972 and 1986 (Vinogradov et al. 1973; Surkov 1977,1986; Surkov et al. 1984). Gamma-ray spectrometers (GRS) were carried on five spacecraft and X-ray flourescence (XRF) instruments were aboard three: the Vega 2 station had both. For convenience, the landing sites are shown in Fig. 1 and the principal results are summarized in Table I.

A major outstanding question for the surface materials sampled by the landers is what part of the rock cycle they represent. The K/U and K/Th ratios of the Venus samples are similar to those of terrestrial volcanics, but match the SNC meteorites and Martian samples even more closely. The K/U ratios are distinct, however, from those of lunar basalts and eucrites. A significant observation is that the K/U ratios are within a factor of 3 of each other. This relative constancy is important because near-surface metamorphic processes apparently have not metasomatically altered the relative abundances of K,U and Th by more than a factor of 3. If this conclusion is correct then the K/U ratios reflect the original bulk composition of the rocks. The ratios imply that the volatile-refractory element inventory of Venus is more like that of Mars and Earth than the Moon or the eucrite parent bodies. Evidence presented below supports this important conclusion.

While the major oxide compositions of the Venusian rocks are broadly basaltic (see below), stronger conclusions are difficult to make with confidence. Three serious uncertainties exist. First, the stated analytical errors are very large, and the disagreement in potassium abundance between the two instruments on Vega 2 suggests additional uncertainty (Table I). An important parameter describing basaltic rocks is the magnesium number, $Mg^* = MgO/(MgO+FeO)$. Typically, $Mg^* = 0.6$ to 0.7 for primitive terrestrial basalts. At face value, Mg^* from the Venera and Vega measurements is certainly consistent with terrestrial basalts, implying that Mg^* for the Venusian mantle is similar to terrestrial peridotite. But to calculate Mg^*, we need ferric as well as ferrous iron, i.e., the Fe_2O_3 concentration in addition to FeO. Given that the basaltic rocks are chemically weathered, the FeO/Fe_2O_3 ratio, even if known, would not constrain Mg^* of the Venusian mantle. We conclude, therefore, that such critical measures of igneous petrogenesis are not adequately constrained by the lander measurements.

A second uncertainty is that the samples analyzed are regolith and not intact bedrock. There is no guarantee that these two are equivalent in composition. Venera 9, 10, 13 and 14 panaramas show subhorizontally layered, slab-like rocks. Materials at the Venera 13 and 14 landing sites have low bearing capacity, low density (1.15–1.5 g cm^{-3}) and high porosity (~50%) (Zolotov and Volkov 1992). Evidence for physical weathering is widespread in the panoramas and includes the presence of detrital slabs, fractures, and fine-grained soils between the slabs and dust deposits (Zolotov and Volkov 1992). The distributed occurrence of wind streaks on Venus indicates that aeolian processes operate widely (Greeley et al. 1992). Basilevsky et al. (1985) concluded that the Venera 13 and 14 surface panoramas show volcaniclastic metasediments (cf., Garvin et al. 1984). Hence the samples may not represent pristine igneous rocks but may have been adulterated to various degrees by physical processes.

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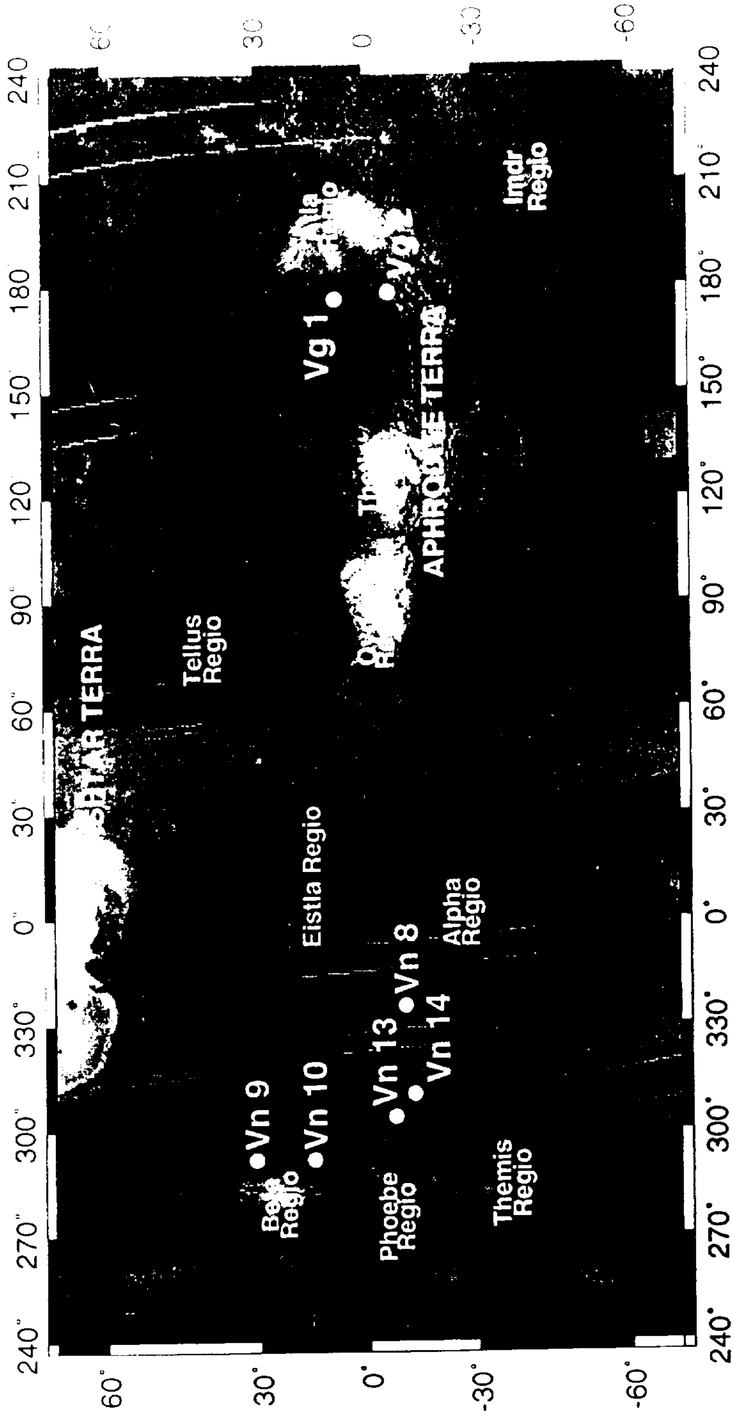

Figure 1. Venera (Vn) and Vega (Vg) landing sites with geochemical measurements. Magellan topographic base in Mercator projection. Nomenclature restricted to place names used in this chapter.

TABLE I

Surface Geochemical Measurements[a]

Constituent	Venera 8	Venera 9	Venera 10	Venera 13	Venera 14	Vega 1	Vega 2
SiO_2	—	—	—	45.1±3.0	48.7±3.6	—	45.6±3.2
TiO_2	—	—	—	1.59±0.45	1.25±0.41	—	0.2±10.1
Al_2O_3	—	—	—	15.8±3.0	17.9±2.6	—	16.0±1.8
FeO	—	—	—	9.3±2.2	8.8±1.8	—	7.74±1.1
MnO	—	—	—	0.2±0.1	0.16±0.08	—	0.14±0.12
MgO	—	—	—	11.4±6.2	8.1±3.3	—	11.5±3.7
CaO	—	—	—	7.1±0.96	10.3±1.2	—	7.5±0.7
K_2O				4.0±0.63	0.2±0.07		0.1±0.08
	4.8±1.5[b]	0.6±0.1[b]	0.4±0.2[b]			0.54±0.27[b]	0.48±0.24[b]
S	—	—	—	0.65±0.4	0.35±0.31	—	1.9±0.6
Cl	—	—	—	< 0.3	< 0.4	—	< 0.3
U (ppm)	2.2±0.7	0.6±0.2	0.5±0.3	—	—	0.64±0.47	0.68±0.38
Th (ppm)	6.5±0.2	3.7±0.4	0.7±0.3	—	—	1.5±1.2	2.0±1.0

[a] wt. %. Table after Barsukov (1992).
[b] K converted to K_2O (K_2O wt. % = 1.21 K wt. %).

The last and perhaps most serious concern is that these sediments are not fragments of igneous rocks but are instead composed of metamorphic rocks. Thermodynamic calculations predict that basalts will react with the hot and caustic atmosphere to produce new secondary minerals (Volkov et al. 1986). Terrestrially weathered and metamorphosed basic to ultramafic rocks not only suffer a change in mineralogy but also in chemical composition. Note that the SO_3 content of the Venusian regolith is elevated, proving that sulfur has been added. What other elements have been added, mobilized, or subtracted?

It is well known from the comparison of metamorphosed and fresh terrestrial basalts and komatiites (Beswick 1982) that elements such as Ca, the alkalis, and Si are mobilized during even modest metamorphic processes. In contrast, elements such as Al or Ti appear to be unaffected by such metamorphic overprints. It is curious, therefore, that the basalts of Venus have low CaO contents and low CaO/Al_2O_3 ratios when compared to their terrestrial counterparts. There is a strong possibility that CaO has been lost relative to Al_2O_3 in the Venusian regolith. If CaO has been redistributed, other elements such as K and Si may have suffered a similar fate. Attempts to use these elements (and others) to constrain petrogenesis may be fraught with incalculable error. But do such processes operate on Venus?

The metasomatic effects observed in terrestrial komatiites and basalts are largely due to the transport of geochemical materials by H_2O-rich fluids. In contrast, Venus rocks in equilibrium with the atmosphere should contain relatively anhydrous minerals given the low H_2O contents (~100 ppm) and high temperatures of the atmosphere (Zolotov and Volkov 1992). But this conclusion does not necessarily apply to the interiors of volcanic flows or ash deposits. Bubble formation of H_2O-rich fluids is hindered on Venus because of the high atmospheric pressures. The solubility of H_2O in basalt melts is roughly 1% by weight at 100 bar, the atmospheric pressure on Venus. This constraint means that basalt melts with smaller amounts of dissolved H_2O (as is true for the vast majority of ocean floor and ocean island basalts on Earth) will not exsolve H_2O. The H_2O in these melts will be trapped in glass (quenched melt) or will exsolve from these melts after extended crystallization brings the melt to the H_2O solubility limit. Under the ambient conditions, the rock should crystallize to greenschist facies mineralogy (chlorite, epidote, micas). During this re-crystallization or during late stage exsolution of volatiles, H_2O-rich fluid is capable of migrating through the rock or ash column, producing just those metasomatic effects alluded to above. It should also be noted that Arvidson et al. (1992) have inferred from radar signatures that over 1 m of erosion or degradation has occurred in the oldest rocks. If correct, the rocks analyzed by Venera and Vega may have originally been in the interior of basalts or ash flows and would have recently been exposed to the atmosphere. Such rocks may not be in equilibrium with the atmosphere.

Nonetheless, a number of compositional inferences have been made assuming that the landers did sample relatively unaltered volcanic rocks representative of the crust. Whether these conclusions can stand close scrutiny

S	—	—	—	0.65±0.4	0.35±0.31	—	1.9±0.6
Cl	—	—	—	< 0.3	< 0.4	—	< 0.3
U (ppm)	2.2±0.7	0.6±0.2	0.5±0.3	—	—	0.64±0.47	0.68±0.38
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[a] wt. %. Table after Barsukov (1992).
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depends on one's prejudices concerning the alteration of the analyzed samples.

Venera 14 is the closest analog to a tholeiite basalt notwithstanding that an analysis for Na_2O is not available. If we assume that Na_2O is 2 wt%, then the analysis sums to about 98%, a reasonable total given the analytical uncertainties. At first glance, the composition of the rock analyzed by Vega 2 appears anomalous in view of its low silica, TiO_2 and K_2O contents. Indeed, McKenzie et al. (1992a) suggest that the composition resembles more a cumulus gabbro than a tholeiite basalt. But if we add about 2 wt% Na_2O to the analysis, subtract the sulfur content and renormalize the analysis to 100%, the SiO_2 content becomes roughly 50% in line with typical tholeiite basalt (Barsukov 1992). The resulting CaO (8.3%) and TiO_2 contents remain on the low side compared to most terrestrial basalts, however. A plutonic origin for this rock cannot be discounted but considering the analytical difficulties this interpretation is highly unconstrained. The analysis of the composition of the rock at Venera 13 is characterized by comparatively low silica but very high K_2O contents. The absence of Na_2O in the analysis is critical to an understanding of the petrogenesis of this rock. The compositions of highly potassic mafic volcanic rocks on terrestrial continental crust typically have $K_2O/Na_2O > 1$ (by weight), whereas ocean-island SiO_2-undersaturated basalts have $K_2O/Na_2O < 1$. K_2O/TiO_2 ratios of 2.5 are more characteristic of ultrapotassic rocks in continental environments, however (see also McKenzie et al. 1992*a*). In detail the high MgO and Al_2O_3 contents and the low CaO contents are very unusual and do not characterize the vast majority of terrestrial rocks.

High K was also found by the Venera 8 GRS, as well as elevated U and Th. Nikolayeva (1990) inferred that this material is intermediate in silica content and stressed that this differs sharply from the Venera 13 sample, in spite of a similar potassium concentration. However, we see no reason to reject the hypothesis that the Venera 8 and Venera 13 sites are similar, as there were no U or Th measurements for the latter, and so it is possible that all three large-ion lithophile elements are partitioned similarly. In either case, an early suggestion that Venera 8 sampled granitic rocks (Vinogradov et al. 1973) now seems unlikely. We refer the reader to Barsukov (1992) and the references therein for more detailed speculations about petrogenesis.

B. Volcanic Morphology

It has long been recognized that the morphology of many volcanic structures on Venus, such as shield volcanoes and very long flows, indicate low-viscosity, effusive eruptions which are in turn characteristic of basalts (Masursky et al. 1980; Barsukov et al. 1986; Head et al. 1992). Such features account for the vast majority of cataloged volcanic landforms on the planet. However, two uncommon classes of structures suggest both extremely low-viscosity lavas and high-viscosity lavas: the canali for the former and the pancake domes and festoon flows for the latter.

The canali are very long (up to several thousand km), narrow, meandering channels found at many locations on Venus. Baker et al. (1992) considered three alternative hypotheses for the operating fluid (neglecting water): komatiitic lava, carbonatite lava, or liquid sulfur. They did not favor komatiite on the basis that its high eruption temperature would also promote rapid cooling and freezing of the lava, before it could produce a long channel. However, they did acknowledge that a combination of low viscosity, high discharge, warm atmospheric conditions, and thermal insulation by a solidified roof (now collapsed) might achieve low cooling rates. Gregg and Greeley (1993) have since shown that such insulating processes might act efficiently on Venus, obviating the problems of rapid cooling. Indeed, lunar rilles, which are comparable to if not as long as the Venusian canali, were formed from mare basalts, lavas which were very hot and fluid. Carbonatites are predicted as near-solidus melts for deep primary melts in the presence of CO_2 (Hess and Head 1990). Sulfur is also reasonably abundant in the Venera XRF analyses. However, Kaula (1993) has pointed out that it may be difficult to concentrate a large amount of low-melting point material. D. McKenzie (personal communication) has suggested that there is no reason to exclude fractionated tholeiite basalt as a channel-forming fluid. Such lava is a common product of differentiation and can travel large distances on Earth as surface flows or in dikes. But to our knowledge, these basalts rarely if ever cut channels into the continental crust, certainly not on the scale observed on Venus and the Moon. In contrast, ultramafic lavas on the Moon, the low and high TiO_2 mare basalts, produce very long rilles. The simplest hypothesis is that silicate magmas were responsible for the channels and these were magnesian in composition with affinities to picrite or komatiite magmas.

The pancake domes—formally called steep-sided domes—have been linked as analogs to terrestrial rhyolite domes and are presently considered the strongest morphological evidence for evolved lavas (Pavri et al. 1992). McKenzie et al. (1992*b*) mathematically modeled these structures as single effusions of a Newtonian fluid and concluded that the inferred high viscosity could be satisfied only by very silica-rich lavas such as dry rhyolite. There are several potential objections to this conclusion, however. Fink et al. (1993) compared the morphology of the steep-sided domes to several terrestrial analogs, and found that high aspect ratios are a feature of episodic rather than continuous effusions. Such episodic eruptions are, however, characteristic of silicic lavas. More recently, Bridges (1995) reported on large, steep-sided basaltic domes on the terrestrial seafloor, which suggests that such morphology can be dominated by eruption conditions rather than composition. Gregg and Fink (1995), scaling from laboratory analogs, concluded that episodic basaltic eruptions or continuous andesitic eruptions could produce the observed morphology. Magma crystallinity is another factor that could influence viscosity (Sakimoto and Zuber 1995). Lastly, Ford (1994) found that the radar-scattering properties of steep-sided domes were similar to those of the surrounding plains.

Even more rare than the pancake domes are the large, steep-sided, ridged, radar-bright "festoon" flows. Moore et al. (1992) mapped one such structure and reported that the distribution and morphology of its deposits suggested local magmatic differentiation. Using a variety of published scaling relations, they concluded that the breadth and thickness of the flow lobes and the large separations of regularly spaced ridges indicated an intermediate to high silica content. Subsequent laboratory analog experiments on flow morphology (Gregg and Fink 1995) and on fold spacing (Gregg et al. 1995) jointly point to an intermediate composition.

In summary, the morphology of the overwhelming majority of volcanic features on Venus indicate basaltic compositions. Although both exotic volatile fluids and highly silicic lavas have been invoked to explain the few unusual features, the possibility that all are silicate melts of ultramafic to intermediate composition cannot be excluded.

C. Linked Geochemistry and Morphology

Basilevsky et al. (1992) and Weitz and Basilevsky (1993) have examined the Venera and Vega landing sites in Magellan images. They found steep-sided domes at the Venera 8 and 13 sites, but not at the others. Although it is unlikely that these spacecraft actually landed on domes, this result does strongly suggest that non-tholeiitic compositions are spatially linked to volcanic structures indicative of somewhat evolved lavas. On a larger scale, however, there is no evident geological correlation in the locations of these sites: Veneras 13 and 14 landed within ~900 km of each other on the eastern flank of Phoebe Regio. This a very heterogeneous region, composed of several tessera blocks separated by plains, with later uplift, rifting, and volcanism associated with the Beta-Themis-Aphrodite disturbances. In contrast, Venera 8, ~2600 km farther to the east from Venera 14, lies in regionally unremarkable plains.

D. Other Approaches

One significant difference between Venus and Earth is their fractional distribution of elevation, or hypsometric curve; Venus' is smooth and unimodal, whereas Earth's is bimodal (Masursky et al. 1980). The split character of Earth's hypsogram is due of course to the distinct mean elevations of continents and ocean basins. However, this does not reflect so strongly the difference in density (composition) between oceanic and continental crust, but instead simply the thicker crust of the continents. The thickness of the oceanic crust is a near-constant 6 to 7 km as a consequence of the mechanics of decompression melting under the same repeating conditions at the mid-ocean ridge (see below). In contrast, the thickness of the continental crust is determined by the feedback between erosion and isostasy; erosion rapidly planes down continental elevations to near the erosional base level (sea level). Isostatic balance with the adjacent oceanic water and rock column (including potential differences between the thickness of oceanic and continental lithospheres) determines the mean thickness of the continents. This thickness has changed only slowly with time, as evinced by the near-constant continental freeboard (Wise 1974; Reymer and Schubert 1984). In the absence of oceans and an erosional base level, Earth's hypsogram might be unimodal (McGill et al. 1982). Alternatively, continental crust could still be magmatically and tectonically thickened and remain high-standing due to the preferential deformation of weaker silicic crust.

Gravity measurements are used to constrain planetary internal density differences, which might be used to search for light, felsic crust. Unfortunately, these differences are too small to detect in the presence of Airy isostatic compensation: the density difference between mafic and felsic crust on Earth (<0.3 Mg m^{-3}) is significantly smaller than the contrast between crust and mantle (>0.5 Mg m^{-3}). Given comparable vertical intervals representative of either the mean crustal thickness or boundary relief on the moho, the latter is likely to dominate any crustal gravity signal. The most likely scenario in which to detect a difference between mafic and felsic crust would be if differences in relief were entirely due to the density contrast between these materials, i.e., an intracrustal Pratt compensation. Such a mechanism is unlikely however; Airy isostatic adjustment is rapid on Earth ($\sim 10^4$ yr), and is controlled mainly by the viscosity of the upper mantle (Cathles 1975). Intracrustal isostatic adjustments could dominate only if the mantle could not respond over the lifetime of the topography, of order 10^8 yr for Venus. This would require an upper-mantle viscosity $\sim 10^4$ times higher than Earth's, which is still unlikely even if the mantle of Venus is drier (Kiefer and Hager 1991). Much of the long-wavelength topography on Venus appears to be dynamically supported (see, e.g., Phillips and Malin 1983), in which case gravity interpretations are even less sensitive to surficial density variations.

E. Summary

Detailed petrological inferences about the crust of Venus are hampered by concerns remaining about the accuracy of the Venera/Vega geochemical measurements and whether or not the surface materials are indeed relatively unaltered volcanic rocks. Nonetheless, mafic to intermediate silica contents are consistent with all geochemical and geomorphological data. This includes the two landing sites that apparently sampled more alkaline rocks and have nearby unusual volcanic structures, and for which no clear regional geological associations are evident. Any notably non-basaltic composition suggests that tertiary differentiation has occurred and been preserved. Although unusual volcanic structures appear to be volumetrically minor, it may be significant that over one-quarter of the landing sites suggest at least some tertiary crustal differentiation.

V. ESTIMATES OF CRUSTAL THICKNESS AND VOLUME

Although geophysical techniques have not been successful in determining crustal density (and hence composition), they have provided reasonable estimates of crustal thickness. This single measurement in turn places strong constraints on models of crustal generation. We therefore review these techniques in detail. There are three approaches: theoretical considerations based on the phase relations, geodynamic models, and gravity-topography relationships.

A. Theoretical Limits

The thickness of a gabbroic (basaltic) crust is limited by $P-T$ stability conditions (Fig. 2). Partially molten crust is not likely to be stable for hundreds of millions of years, and so equilibrium crustal thickness–temperature combinations to the right of the solidus may be ruled out. In addition, gabbroic crust will transform to denser garnet granulite and then eclogite with increasing depth and become gravitationally unstable. If the density increases approximately linearly from $\sim$2.9 Mg m^{-3} to $\sim$3.5 Mg m^{-3} across the granulite field (Namiki and Solomon 1993), then the density of the crust will not exceed the density of the mantle (3.3–3.4 Mg m^{-3}) until it is fairly near the eclogite boundary. These considerations limit the thickness of the crust to somewhere in the range 30 to 100 km, depending on temperature. Note, however, that these boundaries also depend on bulk composition. The selected curves (Ito and Kennedy, 1971) show the largest separation of the granulite and eclogite phase boundaries among the measurements summarized by Ringwood (1975). The full range of compositional variations (shaded region in Fig. 2) then allows the eclogite phase transition to lie practically anywhere between 10 and 100 km. Nonetheless, note also that these boundaries to gabbro stability are substantially shallower than those originally proposed by Anderson (1980).

Two thermal conditions serve as additional constraints on eclogite stability. First, this phase transition is unlikely to limit crustal thickness unless the temperature gradient is $\leq$10 to 15 K km^{-1}, as the crust would melt first. Second, the eclogite phase transition may be kinetically inhibited at low temperatures (Namiki and Solomon 1993; Jull and Arkani-Hamed 1995). The temperature T^* for which this restriction applies over the present surface age of the planet may be roughly quantified by a simple dimensional analysis. The expression for ionic diffusion (Namiki and Solomon 1993, their Eq. 2) has a characteristic time constant $\tau \sim r^2/D$, where r is the grain radius and D is the diffusion constant. Minimum and maximum values for $D(T)$ are given in Eq. (3) of Namiki and Solomon (1993) and a range of r = 1 to 10 mm was also chosen by these authors. A lower bound on T^* for τ = 400 Myr follows by minimizing r and maximizing D, for which we find T^* = 970 K. The maximum blocking temperature that will achieve the same diffusion time (given by the maximum r and minimum D) is 1380 K. This simple dimensional analysis is in good agreement with Fig. 4 of Namiki and Solomon (1993). For comparison, Parmentier and Hess (1992) chose T^* = 870 to 1070 K.

Although the determination of reaction rate has been approached theoretically, quantitative experimental investigation has not followed, primarily because of the complexity of the reaction mechanisms and because of the slow rate of reaction. Observations of natural terrestrial eclogite settings may pro-

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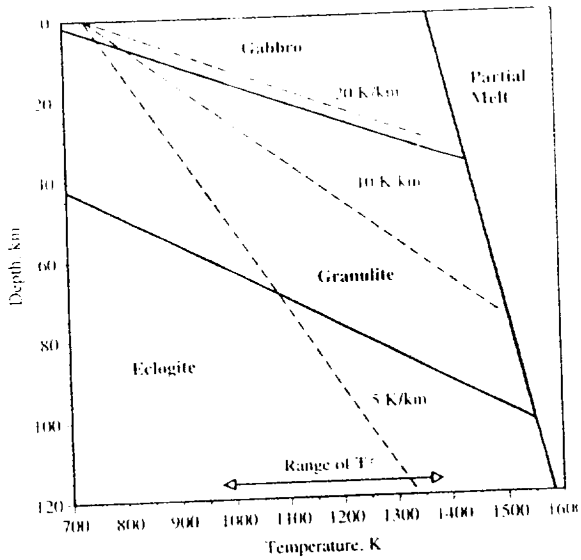

Figure 2. Phase boundaries for gabbro, garnet granulite, eclogite, and the gabbro solidus on Venus (Ito and Kennedy 1971). Shaded region shows range in granulite-eclogite phase boundary due to differences in composition (Ringwood 1975). Representative equilibrium conductive geotherms of 5, 10, and 20 K km^{-1} are shown as dashed lines. Granulite and eclogite phase transitions are inhibited at temperatures below T^* over the age of the present Venus surface, ~400 Myr. The range in T^* and in the positions of the phase boundaries do not allow controls on crustal thickness due to phase changes to be specified with confidence.

vide some insight (Herzog and Hess 1996). Terrestrial eclogite samples yield information about conditions of equilibration through equilibrium mineral compositions, and information about reaction mechanisms can be gleaned if textural features indicating partial reaction can be identified. Eclogite-forming reactions in natural rocks equilibrated at temperatures greater than 750 C usually do not show evidence for fluid infiltration (Rubie 1990). This may indicate lack of free fluid during reaction, or perhaps, expulsion of fluid during the early stages of reaction. Some field evidence shows that eclogite-bearing shear zones alternate on a scale of meters with the dry, unsheared granulites (Austrheim and Griffin 1985). These features demonstrate that granulite production is kinetically easier than eclogite because hydrous fluids, which facilitate rapid reaction, may be removed from the rock during metamorphism of basalt to granulite.

The question dealing with the gabbro-granulite-eclogite transition on Venus may therefore hinge on the question: how much H_2O is contained in the crust? We have argued earlier that volcanic rocks will immobilize H_2O within greenschist facies minerals provided that the H_2O content of the erupted liquids are less than about 1%. This constraint does not apply to plutonic melts where the solubility of H_2O increases with pressure. Because the intrusion/extrusion ratio is certainly greater than one and probably closer to about 10 on Venus (Crisp 1984), it follows that the crust is largely of a plutonic origin. If true, the H_2O content of the crust, particularly the lower crust, should be higher than implied by the relatively dry atmosphere. Studies of ^{40}Ar and H_2O loss are at present very model dependent, and no firm conclusions have been reached as to whether the interior of Venus is very dry (Namiki and Solomon 1995) or has simply not efficiently degassed (Kaula 1990,1993). As the crust passes into the granulite and eclogite regime a series of dehydration reactions ensue, releasing H_2O, enhancing the kinetics of solid state processes, and hence lowering T^*. Whether H_2O content of the Venus crust is sufficient to accelerate the kinetics of the granulite-eclogite transition is unknown. Other lines of reasoning including the apparent strength of the Venus crust point to a relatively dry crust (see below).

This discussion illustrates the strong dependence of crustal stability upon temperature and composition, particularly the H_2O content. If the granulite/eclogite boundary is deep and/or T^* is large, then the eclogite transition might never occur, regardless of crustal thickness. On the other hand, a shallow phase boundary and a low T^* could globally limit the crustal thickness to as little as 10 to 20 km. Intermediate values of all parameters lead to limiting crustal thicknesses ~50 km for $dT/dz > 5$ to 10 K km^{-1}. We conclude that there is insufficient information at present to place a firm limit on crustal thickness due to phase transitions.

B. Geodynamic Models

These techniques seek to understand the forces that deform a planet's surface and the physical parameters that control the rate and style of deformation. In general, these models are responsive to the thickness of the elastic or mechanical lithosphere and not the crust. However, the "yield envelope" concept clearly shows the dependence of lithospheric yield strength upon crustal thickness and temperature gradient (Kohlstedt et al. 1995; chapter by Phillips et al.). Frictional sliding governs the yield strength of the uppermost lithosphere and is independent of rock composition. However, thermally activated creep dominates at greater depths and is a strong function of both temperature and composition. As the crust (even basaltic) contains more weak felsic minerals than the strong ultramafic mantle, increasing the thickness of the crust allows more rapid deformation at a fixed stress, or conversely, a smaller "creep strength" at a specified strain rate. Similarly, creep rates increase and rocks weaken under higher temperature gradients.

Two classes of geodynamic models have been used to estimate the thickness of the crust of Venus. In the first, Grimm and Solomon (1988) modeled the isostatic rebound of impact craters as either an elastic or viscous process. For both models, the topography of craters observed by Venera 15 and 16 constrained the crustal thickness H to <10 to 20 km for thermal gradients dT/dz = 10 to 20 K km^{-1}. In essence, the crustal rocks appeared to be too weak to support topography over geological time scales, and so the models required that strong mantle rocks be placed relatively close to the surface.

The second kind of approach was taken by Zuber (1987) and Banerdt and Golombek (1988). These workers used elastic, viscous, and plastic rheologies to model the characteristic spacings of tectonic structures. In general, the spacing of structures increases with lithospheric thickness. However, in some cases two superimposed wavelengths are apparent, which indicates two scales of strength in the lithosphere. The shorter spacing is controlled by the strong upper portion of the crust. The longer wavelength is thought to arise from the strong upper mantle or the suppression of intermediate wavelengths in the weak lower crust, depending on the model assumptions. In either case, a weak lower crust must separate strong upper crust and strong upper mantle. For such regions of multiple wavelengths, Zuber found H = 5 to 30 km and $dT/dz < 25$ K km^{-1}, whereas Banerdt and Golombek (1988) constrained H to 5 to 15 km and dT/dz to 10 to 15 K km^{-1}.

The crater rebound and tectonic wavelength techniques yielded remarkably similar results, thus mutually reinforcing each other. The conclusion that the crust of Venus was thin, <30 km, was a sharp departure from previous ideas, especially Anderson's (1980) suggestion that the lower uncompressed density of Venus than Earth could be explained by a very thick crust (>100 km).

Two recent developments call for revision of these results. First of course are the Magellan images and altimetry, which can be used to revise the crater isostatic rebound approach by searching for evidence of crater floor uplift and associated fracturing. In addition, various tectonic structures can be studied to update deformation wavelength studies. Second, Mackwell et al. (1995) have performed creep measurements on dry diabase, which may provide a better analog to the basaltic crust of Venus than previous flow laws. They found that the strength of crustal rocks under anhydrous conditions can approach that of the ultramafic mantle. In this case the strength stratification becomes less distinct, and so crustal thickness constraints cannot be derived (Hillgren and Melosh 1995; Brown and Grimm 1996*a*). The overall thickness of the mechanical lithosphere still depends on dT/dz, however. Future revision of the tectonic modeling will determine if the crust and mantle have been sufficiently decoupled during deformation to produce multiple wavelengths of structures and hence constrain crustal thickness. This will depend on both careful inspection of Magellan images to determine if multiple wavelengths are indeed contemporaneous, and the application of the anhydrous crustal flow law.

C. Gravity

The improved orbital geometry and tracking of Magellan have allowed gravity studies to replace geodynamic models as the principal way to infer crustal thickness on Venus. The main problem is to isolate the part of the gravity signal due to crustal thickness variations from other contributions to compensation.

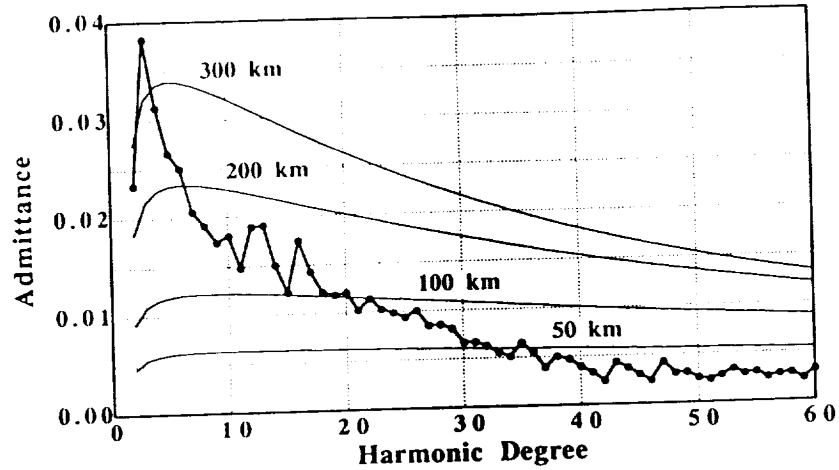

Figure 3. Evidence from gravity for a 20 to 50 km mean crustal thickness. Global spectral admittance from Konopliv and Sjogren (1994), showing short-wavelength asymptote ~25 to 50 km.

Pioneer Venus showed that long-wavelength gravity and topography are highly correlated on Venus, which indicates an apparent depth of isostatic compensation (ADC) of 100 to 200 km (see, e.g., Kiefer et al. 1986). Phillips and Malin (1983) established that classical Airy isostasy could not support topography as any variations in crustal thickness at such great depths would quickly viscously relax and be eliminated (and the topography along with it). We note that this analysis has not been rigorously upheld in the post-Magellan view of Venus, in which the crust is thought to be significantly stronger (see above) and in which thermal gradients may be as low as 5 K km^{-1} in a thermal lithosphere >200 km thick (Sandwell and Schubert 1992; Turcotte 1993). However, simple calculations using the effective viscosity of the crust (Mackwell et al. 1995) and the relaxation time of compensated topography (Solomon et al. 1982) suggest that creep will still be activated at depths comparable to the ADC.

Rejection of crustal isostasy for the long-wavelength topography of Venus led Phillips and Malin (1983) to propose that compensation was due to thermal convection within Venus, and so the "hot-spot" model was born. Smrekar and Phillips (1991) further showed that the individual domal volcanic highlands are the most conspicuous class of deeply compensated features, which further reinforced the hot-spot model and concepts of thermal/dynamic compensation. However, these works and others that examined the role of varying both crustal and thermal lithospheric thickness in order to match gravity and topography (see, e.g., Banerdt 1986; Williams and Gaddis 1991; Grimm and Phillips 1992; Herrick and Phillips 1992) did not explicitly solve for mean crustal thickness, but specified it *a priori*.

Recent gravity models have exploited the higher resolution of the Magellan data to show reliably the spatial and/or spectral variations in compensation depth and in some cases to solve for crustal thickness explicitly. Konopliv and Sjogren (1994) synthesized Magellan line-of-sight spacecraft accelerations into a spherical harmonic model for vertical gravity. Using a spectral admittance technique (the ratio of gravity to topography in the wavenumber domain), they calculated the ADC as a function of wavelength (Fig. 3). The compensation depth is large at long wavelengths, showing the contribution of mantle density variations to the support of topography. However, the short-wavelength ADC has an asymptote at ~25 km, which Konopliv and Sjogren interpreted as the average crustal thickness. They noted that improvements to the gravity models would boost the power at short wavelengths and increase the apparent crustal thickness; however, they claimed that it would be unlikely to exceed 50 km. This result should be interpreted with some additional caution, as mantle thermal and/or compositional compensation could still have an effect across the spectrum. We therefore interpret their 50-km value as an upper limit to the crustal thickness.

Simons et al. (1994) also computed spectral admittances from the spherical harmonic model of Konopliv and Sjogren, but used moving windows to highlight regional variations in compensation. They found short-wavelength ADCs of ~50 km for several plateau highlands, confirming and extending earlier work by Smrekar and Phillips (1991). Kucinskas and Turcotte (1994) used a spatial admittance technique to compute a zero-elevation crustal thickness of 50 km for Ovda and 60 km for Thetis. We believe that the result for Thetis is influenced by deeper, noncrustal compensation (see below) and so take the 50-km value to be representative of the crustal thickness calculated by this method. McKenzie (1994) roughly estimated the crustal thickness at Beta Regio to be ~30 km using the residual gravity from a simple space-domain filter and an estimate of upper crustal extension from radar imagery. However, this hybrid technique requires additional assumptions about extensional geometry, particularly that strain throughout the crust is vertically homogeneous.

Phillips (1994) performed a comprehensive study of the compensation of Atla Regio, a type example of a domal volcanic highland. He solved for the thicknesses of the crust, elastic lithosphere, and thermal lithosphere by minimizing misfits in the wavenumber domain. Phillips' best-fitting mean crustal thickness for the short-wavelength portion of the signal was 30±13 km. The crustal thickness was unbounded for the long-wavelength band. Subsequent work using an improved gravity model (chapter by Phillips et al.) has failed to confirm these results, however.

Grimm (1994) focused on the gravity of four plateau-shaped highland regions. Compensation was partitioned into an upper layer for the crust and a lower layer representing mantle density variations (thermal or chemical). Alpha, Tellus, and Ovda Regiones were found to be well fit in the spatial domain by essentially complete compensation in a crust of thickness 35 to 45 km (Fig. 4). A deep component to the compensation of the fourth highland, Thetis

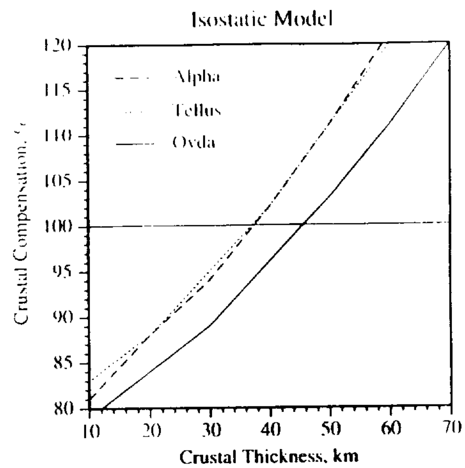

Figure 4. Evidence from gravity for a 20 to 50 km mean crustal thickness. Best-fitting crustal thickness beneath three plateau highlands from Grimm (1994). Isostatic correction to surrounding plains yields ~30 km for all three regions.

Regio, did not allow a unique solution for crustal thickness. Sensitivity studies examining the size of the study area vs the size of the highland showed that such results are biased away from the mean *H* of the study area and towards the crustal thickness existing below the highland. It is then a simple matter to correct the derived crustal thickness to that in the surrounding plains under conditions of Airy isostasy. For ~1 km of average long-wavelength (>1000 km) relief at Alpha and Tellus Regiones and ~2 km of such topography at Ovda Regio, the mean crustal thickness in the surrounding plains for all three regions is ~30 km. Errors and uncertainties allow a range in mean thickness of at most 20 to 40 km.

The full range of mean crustal thicknesses from these studies is 20 to 50 km. We prefer the 30-km best fit from Grimm (1994) because this study explicitly attempted to isolate crustal compensation for several widely spaced tessera, which even competing geological models suggest were formed by thickened crust (Bindschadler et al. 1992; Phillips et al. 1991). Also note that all of the quoted values are estimates of the mean, and not the actual range in crustal thickness: the latter is likely to be comparatively small over most of Venus, as is evident from the planet's smooth hypsogram, considering that long-wavelength, deeply compensated topography contributes substantially. However, several kilometers of relief is associated with some crustally compensated plateau highlands, which require a local thickening of the crust by ~20 to 40 km, i.e., roughly doubling it. The comparatively small area of crustal plateaus implies that the crustal thickness in the plains is close to the global average.

D. Summary: Venusian and Terrestrial Crustal Volumes

Improved gravity data from Magellan have allowed more robust estimates of the mean crustal thickness (~30 km) over previous geodynamic models. It is unlikely that this thickness is globally limited by the gabbro-granulite-eclogite phase transitions. However, the range of compositions and activation temperatures for the phase transition do not allow this hypothesis to be ruled out.

TABLE II

Comparison of Crustal Volumes of Venus and Earth[a]

	Venus		Earth	
	Secondary Crust	Tertiary Crust	Secondary Crust	Tertiary Crust
Present volume (% of planetary volume)	1–2	≪1 ?	0.5[b]	0.5[c]
Recent production rate ($km^3\ yr^{-1}$)	<4[d]	?	20–30[e]	1–6[f]
Time-integrated volume (% of planetary volume)	?[g]	?	~10[h]	~1–2[i]

[a] Rounded to one significant figure.
[b] 61% of surface area occupied by oceanic crust (Turcotte and Schubert 1982) with mean thickness of 6 to 7 km (Reid and Jackson 1981), lower 15 km of 40-km thick continental crust (Christensen and Mooney 1995).
[c] Upper 25 km of continental crust.
[d] Head et al. (1992); Bullock et al. (1993); Strom et al. (1994).
[e] Mean seafloor spreading rate 3.45 $km^2\ yr^{-1}$ (Parsons 1982), 0.3 to 4 $km^3\ yr^{-1}$ combined oceanic and continental basaltic intraplate magmatism (Reymer and Schubert 1984; Crisp 1984; White and McKenzie 1989), mafic portion of arc magmatism 0.3 to 4 $km^3\ yr^{-1}$ (prior references; one-third to one-half of arc magmatism from DeBari and Sleep [1991]).
[f] Reymer and Schubert (1984); Crisp (1984); less mafic portion arc magmatism above.
[g] Depends on presence of recycling and recurrence interval of "catastrophes."
[h] Same rate over 4 Gyr.
[i] Present volume plus recycled sediments, 0.6 $km^3\ yr^{-1}$ (Reymer and Schubert 1984) to 2.5 ± 1.2 $km^3\ yr^{-1}$ (DePaolo 1983), over 4 Gyr.

The range of estimated mean crustal thicknesses reviewed here suggests that the basaltic secondary crust of Venus occupies 1 to 2% of the total planetary volume. This is rather greater than Earth's present basaltic crustal volume, but is significantly smaller than the amount of secondary crust that Earth may have produced by seafloor spreading throughout geologic time (Table II). Earth's time-integrated tertiary crustal production has probably been limited to a few percent or less of its total volume. The uncertainties in the composition of the Venus surface from both chemical and morphological studies (discussed above) preclude a robust estimate of the amount of tertiary crust on Venus.

VI. SPECULATIONS ON CRUSTAL HISTORY

We now review some key inferences about the crust of Venus, and examine how they constrain three alternative models of crustal production and recycling. Because the crust of Venus seems to be predominantly basaltic, much of it appears to have been produced by regional decompression of the mantle. On Earth, this occurs at mid-ocean ridges and hot spots. A combination of horizontal stretching and vertical upwelling emplaces new crust. Seafloor spreading involves infinite lateral strain, and therefore may be described as an environment of horizontal crustal accretion. In contrast, lateral strains are generally modest at hot spots (unless they occur on a spreading center, like Iceland), and so these may be generally characterized as environments of vertical crustal accretion. These two environments may generate broadly similar magmas, but they carry very different implications for the global geology of Venus.

Although basaltic melts are produced by stretching and upwelling on regional scales of ~1000 to 2000 km on Earth (White and McKenzie 1989), the apparently random distribution and relatively pristine morphology of impact craters on Venus indicates a global volcanic flooding event (Schaber et al. 1992). This synchroneity of plains emplacement does not require that the whole crust was generated in a short time, but rather that for some time much less than the crater retention age the entire planetary surface could be reached by flows and/or intrusions from a melt source. This may have taken the form of many plumes or sites of extension. At a minimum, the required extrusive volume is $\sim 10^8$ km^3, given by the thickness required to just obliterate any prior cratering record (a few hundred meters to cover crater rims). Given that intrusions likely outweigh extrusions by an order of magnitude (Crisp 1984), the total volume is probably ~ 100 times larger than individual major continental flood basalt provinces on Earth (Hess 1989; White and McKenzie 1989). The duration of major volcanic floods on Earth is typically only several Myr, so the number of active plumes or extension sites need only be ~10 at any time. Alternatively, plume heads on Venus might be larger than those on Earth by a factor of 2 to 3 if the core-mantle thermal boundary layer is similarly thickened (Herrick and Phillips 1990), which would also lead to ~10 sites.

The comparative uniformity of crustal thickness on Venus is also remarkable. On the Moon, the excavation and isostatic adjustment of impact basins has left many regional variations in crustal thickness of several tens of km (Bratt et al. 1985; Zuber et al. 1994). The hemispheric dichotomy of Mars has preserved comparable early-formed differences in crustal thickness on an even larger scale (Balmino et al. 1982). However, the larger, more active terrestrial planets Venus and Earth have several mechanisms with which to smooth crustal thickness variations. As discussed above, the thicknesses of Earth's oceanic and continental crusts are determined by the mechanics of decompression melting and by isostatic adjustment in the presence of an erosional base level, respectively. As there are no oceans on Venus, there is no fixed erosional isostatic adjustment. Seafloor spreading is therefore the only global process by which to attain uniformity of crustal thickness on Venus. However, the large size of plains-forming lava flows and the presence of long lava channels (Head et al. 1992; Baker et al. 1992) suggests that low-viscosity lavas traveled hundreds or even thousands of km, and lineaments interpreted as dike swarms cover comparable lengths (McKenzie et al. 1992c; Grosfils and Head 1994). Therefore the crust of Venus could have been built up close to true hydrostatic equilibrium. Alternatively, viscous relaxation may have acted to smooth out variations in crustal thickness caused by lateral variations in vertical crustal accretion (Masursky et al. 1980). Both lateral distribution of magma by dike swarms and collapse of topography by solid-state creep are effective on Earth as well.

Lastly, the mean thickness of the crust places some constraints on crustal origins. As just mentioned in the context of terrestrial oceanic crust, the thickness of primary melt and its approximate composition as a function of mantle temperature, thickness of the mechanical lithosphere, and amount of extension is now well understood (see, e.g., McKenzie 1984; McKenzie and Bickle 1988). These ideas have been previously applied to Venus (Sotin et al. 1989; Solomon and Head 1991), but the implications of a thicker crust have not been assessed.

These observations constrain three alternative end-member models for crustal creation and destruction on Venus: (1) horizontal accretion and recycling; (2) vertical accretion with no recycling; and (3) vertical accretion and recycling.

A. Horizontal Accretion and Recycling

Lateral movements of a sufficiently large scale to dominate crustal generation would have been associated with global lithospheric recycling (note that we distinguish lithospheric recycling in general from plate tectonics in particular; other workers have called for periods of "plate tectonics" on Venus, but the specific geometric form of any such lithospheric recycling is unknown). Secondary crust would dominate the crustal production budget, as on Earth. Also by analogy with the Earth, large-scale lateral recycling may have destroyed any primary crust, if such crust ever formed. Tertiary crust may be produced at convergent margins where secondary crust is remelted and therefore should be spatially distributed in some pattern marking former convergent zones.

Arkani-Hamed et al. (1993) have suggested that the present surface of Venus records a permanent end to lithospheric recycling, when the secular decline in heat flow changed the mode of mantle convection from oscillatory to steady. Turcotte (1993) has called for episodes of lithospheric recycling separating periods of one-plate quiescence on Venus. The active mode is brought on by overcooling and thickening of the thermal boundary layer, which subsequently detaches. In fact, Turcotte's one-dimensional model does not specify what the planform of surface movements would be, whether they would be confined to regional deformations associated with sinking lithospheric diapirs and the associated warm upwellings filling them, or whether large-scale lateral movements would truly develop. Head et al. (1994) have pointed out these same issues of horizontal scale in the context of the combined chemical/thermal boundary layer detachment model of Parmentier and Hess (1992). However, this latter model relies explicitly on vertical crustal accretion and will be discussed below.

The parameters required to produce 20 to 50 km of melt at spreading centers are fairly straightforward to calculate. We closely follow McKenzie's (1984) formulation and adopt the dry solidus curve (D16) and an entropy change during melting of 360 J kg^{-1} K^{-1} given in that paper. Upwelling mantle follows an adiabat defined by a specified value of the mantle potential temperature T_p (defined earlier), and this trajectory changes where the solidus is crossed. The total melt thickness is determined by integrating the melt weight fraction as a function of depth and converting to porosity. For infinite extension at a spreading center, there are no other major parameters; at intraplate hot spots, the thickness of the lithosphere and the amount of extension are also considered (see below). Several other factors that could affect the net crustal thickness are ignored by this model, including mechanisms of melt access to the surface, the role of a rigid lid, and equilibrium vs fractional melting.

The heavy line plotted in Fig. 5 agrees with previous work and illustrates the temperature dependence of crustal thickness produced by adiabatic decompression. The nominal 6 to 7 km thick terrestrial oceanic crust is produced by infinite extension of mantle at T_p~1600 K (McKenzie and Bickle 1988; note that small differences of some tens of Kelvins exist between the referenced study and the present work due to differences in the solidus relation and because the gravity of Venus is used throughout). If the mantle of Venus were only 100 K hotter than Earth's, as predicted from parameterized convection theory (Kaula and Phillips 1981; Stevenson et al. 1983), then ~15 km of melt would be produced (Sotin et al. 1989). However, a 30-km mean crustal thickness on Venus requires T_p = 1800 K, about 100 K hotter than expected. The error bounds on mean crustal thickness allow a range of T_p = 1700 to 1900 K. As a rule of thumb, the crustal thickness produced by lithospheric divergence approximately doubles with each 100 K increment in T_p.

The compositional implications of this model agree well with gross volcanic morphology and are marginally consistent with lander geochemical data. The parameterized approach to predicting melt composition of McKenzie and Bickle (1988) suggests a range in bulk MgO of ~17±4%, i.e., picritic to komatiitic. Certainly the low-viscosity lavas inferred from volcanic morphologies are consistent with highly mafic compositions (see above). The mean MgO measured by Venera 13, 14 and Vega 2 is 10±5%. Given the uncertainties in the experimental methods, in possible alteration of the surface rocks, and in the theory itself, we cannot rule out the possibility of more mafic compositions in agreement with this model.

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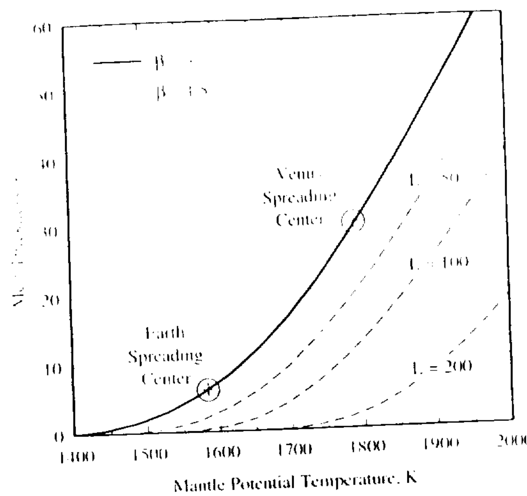

Figure 5. Basaltic crustal generation on Venus. For horizontal accretion at spreading centers ($\beta = \infty$), the best estimate of mean crustal thickness of Venus requires a mantle ~200 K hotter than Earth's. However, the depth at which melting begins must also be much larger, suggesting that spreading centers on Venus must be actively driven by hot plumes in this scenario. Alternatively, crust could be generated in multiple epsiodes of vertical accretion; melt thicknesses of several km at a time can be generated from modest thinning (β = 1.5) of mechanical lithospheres L initially 50 to 100 km thick at mantle potential temperatures only 100 K hotter than Earth.

The larger value of T_p implies not only a hotter mantle, but a greater depth at which melting begins (approximately 140 km at T_p = 1800 K). Such deep melting is indicative of hot jets or mantle plumes. Indeed, McKenzie and Bickle (1988) quote H~30 km and T_p~1800 K as characteristic of terrestrial plumes producing oceanic plateaus such as Iceland. Presumably, such plumes on Venus could be even hotter and form more melt (Fig. 6).

A potential flaw in this end-member model is that it requires *all* of the crust of Venus to be produced by plumes. All divergent margins would have to be underlain by hot upwelling sheets. As divergent margins on Earth are now understood to be the result of passive, or sink-driven, extension rather than as active upwellings (see, e.g., Forsyth and Uyeda 1975), this scenario seems unlikely for Venus.

A second problem for all models involving large-scale lateral movement is the lack of a globally interconnected network of structures indicative of even "fossil" lithospheric recycling (Solomon et al. 1992). Herrick (1994) has argued that the lock-up of lithospheric recycling would briefly raise upper mantle temperatures sufficiently to cause a global melting event and hence bury most pre-existing structures. We consider this mechanism still somewhat *ad hoc*, but there is no better explanation if large-scale lithospheric recycling or "plate tectonics" is invoked, such as in the models of Arkani-Hamed et

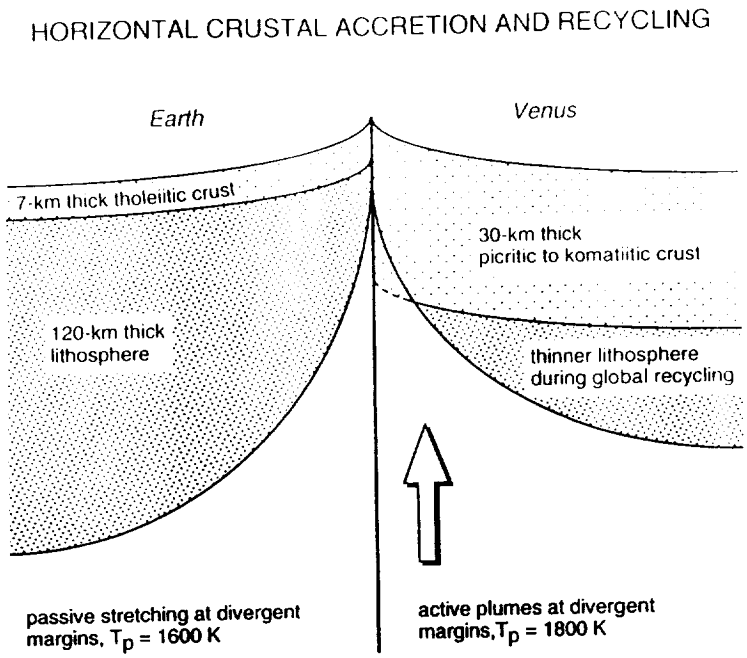

Figure 6. Crustal generation on Venus by horizontal accretion and comparison to Earth. Horizontal model implies crustal recycling.

al. (1993) and Turcotte (1993). We note that statistically significant lateral gradients in crater density diagnostic of horizontal recycling are not likely to be found if recycling was geologically rapid. In other words, if most lithosphere is recycled within 100 Myr, as on Earth, then there will not be a meaningful age gradient on a surface that is 400 Myr old.

B. Vertical Accretion With No Recycling

This end-member model represents another extreme view of the crust of Venus, one broadly more analogous to the Moon or Mars than Earth but which allows episodic global resurfacing (Fig. 7). In this scenario, hotspots formed by mantle plumes and/or modest lithospheric stretching are the agents of secondary crustal production. Tertiary crust forms by remelting of secondary crust, also at hotspots. Without substantial crustal recycling, there is nothing to eliminate a primary crust, and so we must invoke Taylor's (1989) arguments that Earth and Venus never formed sufficient quantities of this crust to stabilize and preserve it. We view doubtfully the possibility that primary crust could be buried and embedded in the secondary crust; a primary crust tens of km thick could be covered by only a veneer of basalt, which would leave Venus a thinly disguised Moon.

McKenzie and Bickle (1988) have examined the volume and composition of melt during lithospheric stretching as functions of the initial thickness of the

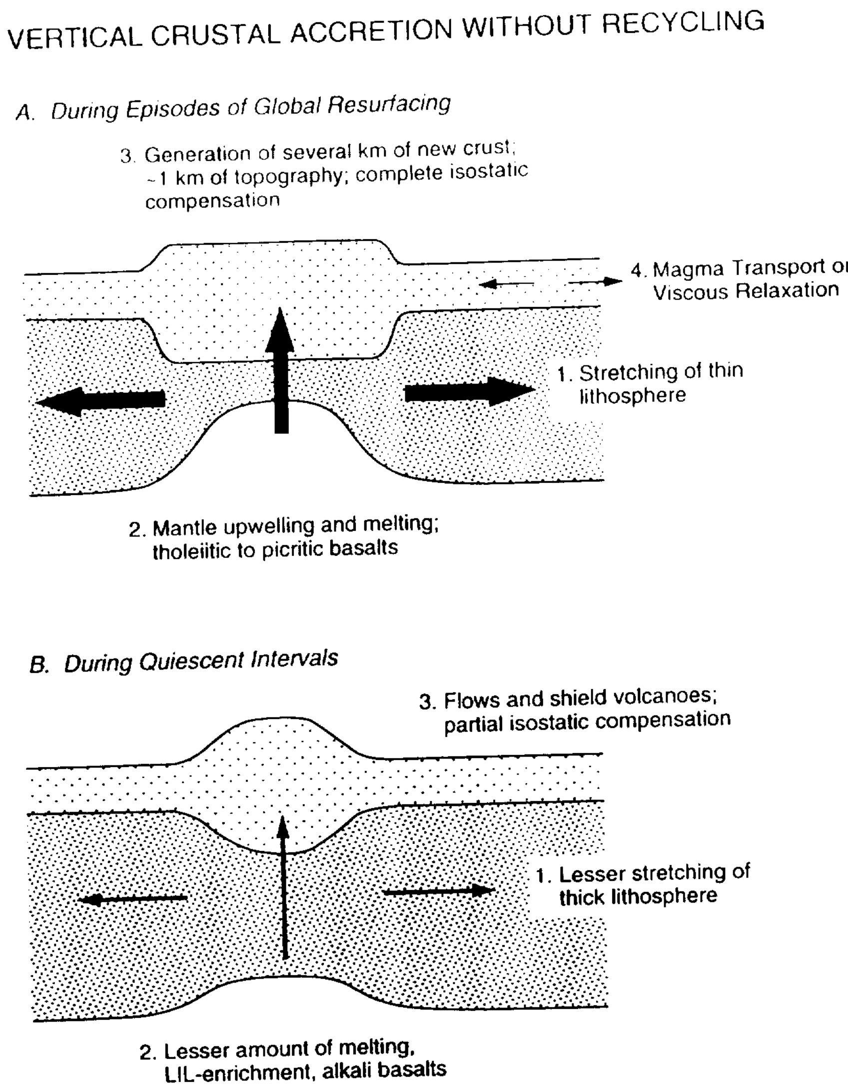

Figure 7. Crustal generation on Venus by vertical accretion, but without recycling.

mechanical lithosphere, the ratio of final to initial surface area β (also given by strain+1), and T_p. Melt production increases with β, attaining a maximum when upwelling approaches the surface at a divergent margin ($\beta = \infty$). A thicker mechanical lithosphere results in lesser melt production.

On Earth, intraplate extension is commonly in the range $\beta = 1.2$ to 2, the larger values being typical of distributed stretching of the Basin and Range (McKenzie and Bickle 1988). Magee and Head (1995) have estimated a mean

$\beta \sim 1.2$ for several rift zones on Venus. This apparent stretching factor is a lower bound, as it has not been corrected for infilling due to rift-generated volcanism (McKenzie and Bickle 1988). If up to several km of melt were added to a crust already a few tens of km thick, then the true values of β are likely to lie in the range 1.2 to 2.0. We adopt $\beta = 1.5$ as exemplary (if not fully representative) of recent rifting on Venus.

The amount of melt generated at $\beta = 1.5$ as a function of the initial thickness of the mechanical lithosphere L is shown as a set of dashed lines in Fig. 5. The present initial thickness of the mechanical lithosphere prior to stretching could be 100 to 200 km (Turcotte 1993; Brown and Grimm 1996*b*). In this case, passive rifting at the nominal $T_p \sim 1700$ K for Venus will generate little or no melt. If T_p is raised 100 to 200 K due to upwelling plumes, several km or more of melt may be produced (note that the melt thicknesses over plumes are likely overestimated for the vertical recycling scenarios as there is no rigid lid in the model). In the upper limits of the melting model, episodes of global plume activity alone may be sufficient to resurface Venus. Several such events, say, one every several hundred million years, could vertically accrete the crust to its present thickness. In the lower limit, minor melt production might characterize Venus during quiescent periods. These conditions are broadly analogous to impingement of mantle plumes upon the terrestrial continental lithosphere, and therefore alkali basalts are expected to be included among the primary melts together with other primitive basaltic to picritic magmas (Fig. 7B). Alternatively, passive extension would dominate if global resurfacing was driven by lithospheric foundering (Parmentier and Hess 1992; Turcotte 1993). The "diapir" mode of sinking would be favored under a vertical crustal accretion model. Kaula (1993) has argued that the high viscosity of the upper mantle inferred from gravity data would lead to a predominance of regional- rather than global-scale mantle movements: a "distributed" convection. He further suggested that a more distributed upwelling of heat would lead to smaller melt production and a lower crustal production rate, obviating the need for crustal recycling.

If the mean thickness of the mechanical lithosphere was significantly less during global resurfacing, say $L < 100$ km, then extension events at $T_p = 1700$ K can each generate several km of melt (Fig. 7A). In comparision to terrestrial mid-ocean ridges, the hotter mantle temperature of Venus will partly offset the smaller amount of stretching, and so the comparatively shallow initiation of melting and similar net melt production should lead to tholeiitic to picritic compositions. A recurrence interval of several hundred million years would also be necessary to generate the full thickness of the Venusian crust.

Impingement of multiple plumes upon pre-existing crust, whether during the same global resurfacing event or in successive ones, may lead to remelting and the production of tertiary crust. Incompatible elements should be the first liberated by small degrees of partial melting, thus enhancing K, U, and Th in the melt products. In the vertical accretion model, tertiary crust should be randomly distributed across the planet and not confined to zones marking prior convergent margins. Crustal remelting could be fairly common, but might not proceed to high degrees of differentiation, if the lander geochemical measurements are representative.

Lastly, this model requires some mechanism to smooth out crustal thickness variations produced by a presumably arbitrary distribution of plumes during multiple crustal accretion events. As discussed above, this could be accomplished by lateral transport of magma by dikes, sills, and low-viscosity flows during crustal accretion, or subsequently by viscous relaxation. One problem with the latter idea is that there are no obviously relaxed impact craters, some of which should date to a time of high heat flow just after global resurfacing (Brown and Grimm 1996*a*). These objections might be removed by allowing impact craters to relax into an isostatic state, from which subsequent rebound slowed, or simply to invoke resurfacing the crater floors by lava. Nonetheless, lateral magma transport seems to be the simpler hypothesis.

C. Vertical Accretion With Recycling

This last end-member model differs from the previous model only in that large amounts of the crust are periodically stripped away and recycled into the mantle (Fig. 8). Primary crust may have been recycled. Secondary crust is generated by regional lithospheric stretching and tertiary crust by remelting of older secondary crust. Vertical recycling is accomplished either by detachment of eclogite or by lithospheric delamination.

In the model of Parmentier and Hess (1992), crustal thickness is controlled by both the depth of the eclogite phase transition and by a kinetic blocking temperature. Both the crust and buoyant melt residuum layer thicken with time, but the growing residuum layer further isolates the crust from the convecting mantle and lowers the thermal gradient, inhibiting the phase transition and/or solid-state creep in the lower crust. When the residuum itself is cool and dense enough to detach, the sudden heat pulse to the crust allows the eclogite to sink. Crustal recycling must occur in this model, as a sufficient thickness of crust must be produced so that its complementary residuum can thicken to the point of cooling and subsequent sinking. The eclogite phase boundary produces a natural horizon controlling crustal thickness variations, maintaining a comparatively uniform crustal thickness over time (Fig. 8A).

The predicted range of crustal thickness in the model of Parmentier and Hess (1992) is greater than that inferred from gravity data. In the model, crustal thickness oscillates between about 50 and 100 km, whereas the inferred thickness of the crust is only 20 to 50 km. Furthermore, the thinnest crust in the model occurs just after catastrophic overturn; the higher end of the range would be more appropriate at present. Recognizing that this preliminary model was intended only to illustrate a new concept, three alternatives exist to refine it and reconcile this disparity. First, the melting relation used was simplified: by not accounting for the increase in solidus temperature with melting, it over-estimates the amount of melt. Second, the granulite and

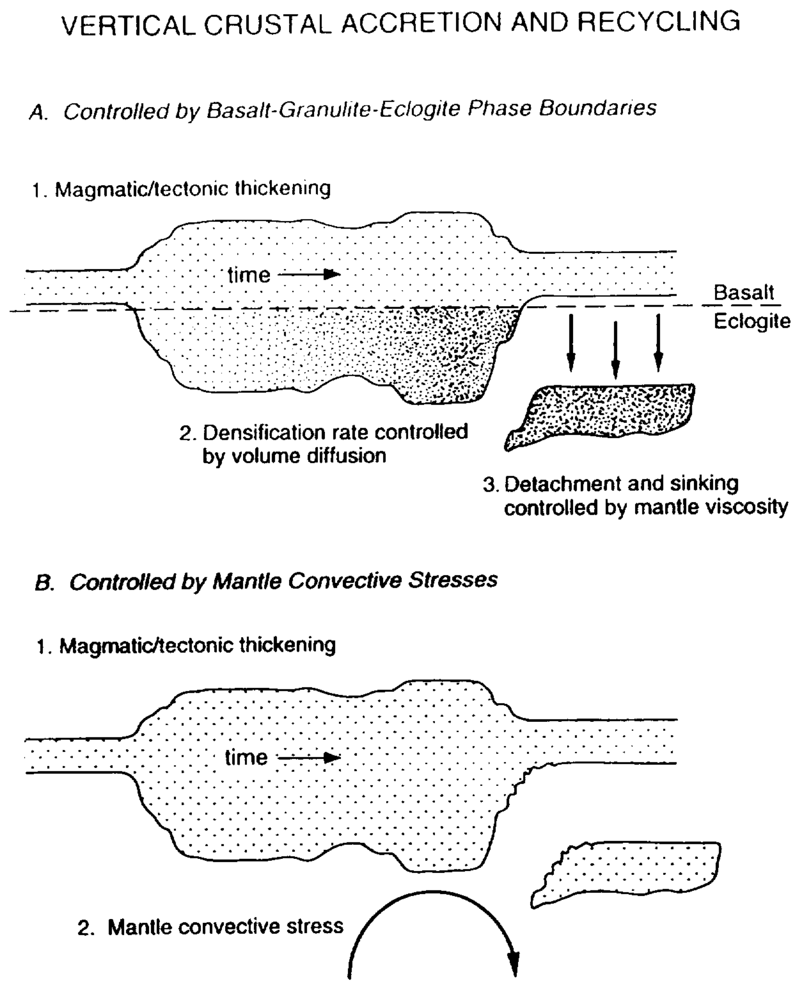

Figure 8. Vertical accretion and recycling models for the crust of Venus.

eclogite phase boundaries could be much shallower than adopted in the model. As discussed above, the eclogite boundary in particular depends strongly on composition, and could lie anywhere in the range ~10 to 100 km. Third, in contrast to the basic model, the granulite and eclogite phases may have little to do with crustal recycling, which may be more strongly controlled by detachment of the chemical and thermal boundary layers.

Lithospheric delamination can include recycling of both chemical and thermal boundary layers. Detachment of dense mantle may place fresh, hot mantle in direct contact with the crust, causing a drastic reduction in viscosity and allowing rapid crustal flow. If sufficiently thinned and stretched by mantle flow, even buoyant crust may be recycled (Fig. 8B; Lenardic et al. 1993).

D. Summary

Of the three end-member models, only horizontal accretion and recycling can be ruled out, for several reasons. Hot plumes must actively drive global lithospheric spreading, in contrast to the passive or sink-driven divergent margins of Earth. The predicted crustal composition may be too mafic. Lastly, a "fossil" plate boundary network is predicted but not observed. It is also difficult to incorporate the horizontal model into a hybrid scenario, unless the dominant mode of crustal generation has changed over geologic time on Venus. For example, 15 to 20 km of crust could have been formed during a prior epoch of lithospheric recycling at T_p = 1700 K. During one or more episodes of more recent global resurfacing, vertical accretion could contribute the remaining 10 to 15 km of crust. Herrick's (1994) model is derived in the limit of a thick crust produced by horizontal accretion and only a thin veneer of vertically accreted volcanics. Unfortunately, such hypotheses are likely to remain untestable.

If, however, a uniformitarian perspective is adopted over the geologic history of Venus—where individual "catastrophic" episodes may be considered repetitions of a uniformitarian theme—then vertical crustal accretion is likely to have dominated. In this case, the "tie line" between the two remaining models depends only on the amount of crustal recycling. Due to the uncertainty in the position of the eclogite phase boundary, we cannot determine whether this transition is an important contributor to crustal recycling, and hence what role, if any, must be assumed by mantle convective flow in stripping the lower crust. There is no evidence at present to determine whether or not crustal recycling has occurred.

VII. CONTEMPORARY CRUSTAL FORMATION

A number of workers have attempted to estimate the recent rate of crustal formation on Venus. Pre-Magellan estimates, based on the rate of impact crater obliteration by lava infilling (Grimm and Solomon 1987) and equilibrium of SO_2 reaction with volcanic rocks (Fegley and Prinn 1989), allowed an extrusive flux up to 1 to 2 km^3 yr^{-1}. This corresponds to 5 to 20 km^3 yr^{-1} of total crustal production if 5 to 10% of the magma is erupted (Crisp 1984).

The Magellan images have provided much more stringent constraints, particularly in the pristine morphology and random distribution of impact craters. Bullock et al. (1993) found that these data could be matched in Monte Carlo resurfacing simulations only if the post-catastrophe volcanic flux was <0.4 km^{3}yr^{-1}. The global reconnaissance mapping of Price and Suppe (1994) revealed that about 15% of the planet represented by volcanoes, flows, and coronae had a mean crater retention age of <100 Myr, or one-third to one-quarter of global value. This translates to a mean resurfacing rate of ~0.4 $km^2 yr^{-1}$ (assuming the crater retention age is one-half of the maximum age), which in turn implies an extrusive flux <0.4 $km^3 yr^{-1}$ for a maximum resurfacing depth of 1 km. Head et al. (1992) suggested that the flux necessary to form the estimated volumes of individual flows and edifices was <0.1 $km^3 yr^{-1}$. Strom et al. (1994) favored a comparable upper limit to the extrusive flux, 0.15 $km^3 yr^{-1}$, in the epoch following global resurfacing. However, these workers further suggested that preserved impact "splotches" could limit the resurfacing rate to as little as 0.01 $km^3 yr^{-1}$. Using a maximum volcanic flux of 0.4 $km^3 yr^{-1}$, the upper limit to the total crustal production rate is 4 $km^3 yr^{-1}$; however, the net magmatic flux based on other estimates could be less than 0.5 $km^3 yr^{-1}$.

These limits to the recent rate of crustal production on Venus may be compared to the Earth (Table II) and to model predictions. The rate of crustal accretion at Earth's divergent margins has been well constrained by the isochron patterns of the seafloor to 21 to 24 $km^3 yr^{-1}$ since the early Mesozoic (Parsons 1982). Estimates of the rates of continental, arc, and oceanic intraplate magmatism are subject to much greater uncertainty, however. The contribution of arc magmatism has been computed to be anywhere from 1.1 $km^3 yr^{-1}$ (Reymer and Schubert 1984) to 8.6 $km^3 yr^{-1}$ (Crisp 1984). As the lower one-third to one-half of the arc is mafic (DeBari and Sleep 1991), these volumes may be crudely partitioned into secondary and tertiary crust. Estimates for oceanic intraplate magmatism and continental magmatism are also variable, 0.2 to 2.4 $km^3 yr^{-1}$ and 0.1 to 1.6 $km^3 yr^{-1}$, respectively (Crisp 1984; Reymer and Schubert 1984). White and McKenzie (1989) assessed the continental magmatic rate, which is dominated by flood basalts, at 0.4 $km^3 yr^{-1}$. All told, the rate of intraplate basaltic magmatism on Earth lies in the range 0.3 to 4.0 $km^3 yr^{-1}$.

This range for the intraplate flux of basaltic magma on Earth corresponds well to the estimated upper limits to recent crustal production rates on Venus; the total contemporary crustal production rate on Venus therefore is comparable to or less than the rate of intraplate magmatism on Earth. This conclusion had been suggested prior to Magellan by Phillips et al. (1991) and from the first global volcanic inventory from Magellan by Head et al. (1992). These fluxes may represent the similar control on the amount of melt generated by mantle plumes and/or lithospheric stretching in both intraplate and one-plate settings.

These estimates of the crustal production rate on Venus are significantly lower than the rates predicted by Parmentier and Hess (1992). In their model, crustal growth occurs at rates of 5 to 50 $km^3 yr^{-1}$, depending on the time within the cycle of episodic accretion and detachment. This implies 0.5 to 10 $km^3 yr^{-1}$ of volcanism, which lies beyond the range accepted from resurfacing studies. As mentioned above, this model presently over estimates the melt production; a lower rate of magmatism predicted by the model might also be consistent with a thinner crust. The models of Turcotte (1993) and Arkani-Hamed et al. (1993) make no specific predictions about rates of contemporary crustal formation, but a terrestrial intraplate analog may be a likely upper bound.

VIII. CONCLUSION

Both geochemical and geomorphological data indicate that the crust of Venus is overwhelmingly basaltic, forming a volume that at present exceeds that of terrestrial oceanic crust and underplated basaltic rocks. Processes of secondary crustal generation thus appear to dominate the present surface of Venus more than known for any other planet. Some data have been used by other workers to indicate strongly non-basaltic compositions: the incompatible-element-enriched Venera 8 and 13 surface materials, the canali, the pancake domes, and the festoon flows. In contrast, we believe that all of these data are consistent with basaltic to intermediate compositions, or at least that such compositions cannot be ruled out at present. In particular, rhyolites and non-silicate fluids are nowhere required on the surface of Venus. Nonetheless, the unusual compositions and volcanic structures probably do indicate that some tertiary crustal differentiation has occurred. Unfortunately, further details on the volume and composition of tertiary crust still remain unanswered.

Analyses of Magellan gravity data give a range of 20 to 50 km for the mean crustal thickness; we suggest a value of 30 km. The planet's comparatively smooth hypsometry and the deep compensation of long-wavelength topography further indicate that this thickness is fairly uniform; significant thickening is present only beneath the large tessera plateau highlands and some minor thinning may exist beneath the domal volcanic highlands. Global control of crustal uniformity may be imposed through accretion at divergent margins, i.e., during a prior epoch of lithospheric recycling on Venus. However, the required mantle temperature is too high; even if mantle plumes lined divergent boundaries, temperature heterogeneities would probably lead to significant variations in crustal thickness. In addition, the crust would be more mafic than observed and a late *ad hoc* burst of global volcanism is necessary to completely bury structures diagnostic of lithospheric recycling.

Vertical crustal accretion is a more likely alternative. Major episodes of melting, crustal generation, and resurfacing occur in response to either upwelling mantle plumes or regional stretching, the latter linked to downwelling of an old lithosphere. Tholeiitic to picritic basalts are expected under these conditions. Heterogeneity in crustal thickness would be minimized by intrusive lateral transport by dikes and sills and extrusive flow by low-viscosity lavas. Minor addition to the crust takes place during quiescent periods at rates up to a few tenths of a cubic km per year, comparable to or less than the intraplate magmatic rate on Earth. Because of the thick lithosphere and modest stretching at present, alkali basalts are produced, analogous to terrestrial continental rifts.

We are unable to answer definitively whether significant crustal recycling has occured on Venus. This is partly due to the uncertainties in the nature of tertiary crust described above. It also follows from the view that densification due to the garnet granulite and eclogite phase transitions are the principal agents of vertical crustal recycling. Given evidence presented here that the crust may be too thin to be limited by these phase boundaries, such recycling is doubtful, or must be accomplished by other mechanisms.

Based on the inferences about its crustal history presented in this chapter, Venus has had periods of substantial surface mobility compared to one-plate planets, resulting in sufficient lithospheric stretching to produce batches of crust several kilometers or more in thickness at a time. The mechanisms of one or more such global catastrophes are the subjects of ongoing study. Nonetheless, it appears that a dry, stiff mantle did not allow the lithosphere sufficient freedom to produce an interconnected recycling network. Venus may be more of a "super Mars" and an ever more distant relative of Earth.

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